1973

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1973 PROGRESS REPORT

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US/IBP DESERT BIOME
RESEARCH MEMORANDUM 74-45

Reprint from Reports of 1973 Progress, Volume 3: Process Studies
Abiotic Section, pp. 15-28

1973 Proposal No. 2.3.5.2.

MAY, 1974

Citation format: Author(s). 1974. Title.
US/IBP Desert Biome Res. Memo. 74-45. 14 pp.

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ABSTRACT

Hydraulic conductivity and soil water diffusivity have been measured for "Rock Valley" gravelly loamy sand and Tubac and Rillito gravelly sandy loams over a soil water pressure range of -0.05 to -50 bars, using a transient outflow method.

A secondary objective of this study was to determine whether moisture transmission properties of these stony soils could be evaluated using samples of the same soils in which the stony fraction (>2 mm) had been excluded. If expressed as a function of soil water pressure, hydraulic conductivity values were similar whether or not stones were present. When K was expressed as a function of volumetric water content (Θ), conductivities were higher for a given water content when stones were present. A simple correction of water contents of stone-free samples, based on the stone volume of each soil, adequately accounted for differences observed when water contents were computed on a total volume basis.

Moisture movement induced by thermal gradients in sealed soil columns was studied under steady-state conditions. For Tubac and Rillito soils, water contents in the soil columns remained unchanged during the experiments because the initial water contents were too high. In studies with "Rock Valley" soil, thermal moisture diffusivity L_{wq} decreased from 14.0 to 1.3 x 10^{-3} $cm^2 hr^{-1} deg^{-1}$ as Θ decreased from 0.14 to 0.08, indicating that much of the moisture movement from hot to cold regions probably occurred in the liquid phase. However, the values of L_{wq} were scattered and the influence of stones on this coefficient could not be determined from the experimental data.

A qualitative evaluation was made of the role larger stones may play in the water economy of desert environments by providing water condensation surfaces on their undersides. Condensation of water vapor could occur whenever the temperature beneath a stone is lower than the dew point temperature. Two experiments -- one involving a buried stone, the other a surface stone -- were set up to measure the temperature distributions under and around stones submitted to a diurnal heat wave. Temperatures were monitored both in air dry and moist soil. Only when a stone was placed on the surface of air dry soil were temperatures lower than in the surrounding soil. A maximum difference of 7 C was attained a few hours after heating began. During the cooling period, the trend was reversed. When soil is sufficiently dry so that water moves mainly in the vapor phase, condensation could occur during the early part of the day. Whether such moisture accumulation is of significance to desert fauna and flora warrants further investigation.

INTRODUCTION

The original purpose of this project was to develop a predictive model for estimating patterns of water withdrawal from soil as a function of depth and time. Before such a model can be tested and evaluated, soil water transport coefficients associated with each gradient component of flow must be determined as functions of depth and water content.

During 1972, a method was selected to evaluate hydraulic conductivity and soil water diffusivity and extended to cover the soil water pressure range of -0.05 to -50 bars. Values of these coefficients were determined for Sonoita sandy loam from the Santa Rita Experimental Range in Tucson Arizona (Letey et al., 1973). Similar work was conducted in 1973 for two soils from the Silverbell Validation Site in Arizona and for one soil from the Rock Valley Validation Site in Nevada. These soils contained large amounts of coarse fragments. Their influence on hydraulic conductivity was assessed.

Thermal moisture diffusivities were determined for the same three soils. Similar work will be carried out for soils sampled at the Jornada Validation Site in New Mexico. Particular emphasis will be placed on moisture movement under thermal gradients.

OBJECTIVES

The overall objective of this study was to develop and test under field conditions a theoretical model for predicting water withdrawal patterns from soil as a function of depth and time in the presence or absence of plant roots under desert conditions. The objectives given in the project proposal have been revised as a result of discussions with Dr. John Hanks to include basic soil water transfer properties as determined in the laboratory. It was concluded that these functions were not well known for the soils of the Desert Biome validation sites. Sonoita sandy loam from the Santa Rita Experimental Range in Tucson, Arizona, was studied in 1972. The specific objectives of the research conducted in 1973 were:

1. To determine soil moisture characteristic curves for soils from Silverbell and Rock Valley Validation Sites.
2. To evaluate hydraulic conductivity and soil water diffusivity as functions of soil water pressure and water content in the relatively dry range.
3. To determine whether the presence of large amounts of stones in these soils could be accounted for by correcting measurements obtained on stone-free samples.
4. To evaluate thermal moisture diffusivity on these soils.
5. To begin work on temperature distributions under large stones submitted to a diurnal heat wave.

METHODS

The material contained in the Methods, Results and Discussion sections are portions of a dissertation stemming from this study (Mehuys, 1973). Copies may be obtained from the author.

Soil Sampling

Samples of "Rock Valley" (placed between quotes to indicate it is not a recognized soil series name) gravelly loamy sand were dug from the first 50 cm of the profile and placed in 5-gallon containers for shipment to the laboratory. Loose samples of Tubac gravelly sandy loam and Rillito gravelly sandy loam (Silverbell, Arizona) were collected in a similar fashion but to a depth of 25 cm. Undisturbed cores were impossible to take owing to the stoniness of these soils. Field bulk density was determined *in situ* following the procedure known as the rubber balloon, or excavation, method (Blake, 1965).

Laboratory Manipulations of the Samples

The samples were dry-sieved into separates < 2 mm, 2-4 mm, 4-9.51 mm, and >9.51 mm. Each size fraction was oven dried, then weighed to determine its percentage on a total weight basis. The stone density was determined by dividing weight of a subsample of stones by its volume obtained by water displacement. The soil material < 2 mm in size was analyzed for particle size distribution following the hydrometer method. Results are reported in Table 1. The percentage of sand, silt, and clay were computed on the basis of soil alone (stones excluded). The electrical conductivity (EC_e) of the saturation extract and the saturated water content on a dry weight basis are given in Table 2 together with bulk density data. The bulk density of the soil fraction only, ϱ_b, was calculated from an equation based on the following reasoning.

Consider a unit volume of soil containing stones whose total weight is n grams. If the ratio of the weight of soil < 2 mm in diameter to the total weight of the sample is W, then the unit volume of soil can be represented by the expression:

$$\text{unit volume of soil} = \frac{n}{\varrho_T} = \frac{Wn}{\varrho_b} + \frac{(1-W)n}{\varrho_s} \qquad (1)$$

where ϱ_s is the density of the stone fraction of the sample and ϱ_T is the field bulk density, or total bulk density, i.e., stones included. Solving the second and third terms of (1) for the second and third terms of (1) for ϱ_b yields:

$$\varrho_b = \frac{\varrho_s \ \varrho_T \ W}{\varrho_s - \varrho_T (1-W)} \qquad (2)$$

Values of W for each soil may be found in Table 1.

Hydraulic Conductivity

The purpose of these experiments was two-fold: (1) to evaluate hydraulic conductivity and soil water diffusivity over a relatively wide soil water content range, and (2) to evaluate these coefficients for the natural stony soils from values obtained on soil samples in which stones were excluded.

Air dry soil material was packed into acrylic plastic columns, 10 cm in diameter by 30 cm in length, with a piece of acrylic plastic taped at one end to form a soil container. To achieve a packing density similar to field bulk density, a pre-weighed amount of material to fill a 2 cm section of the column was added at a time. After each addition, the material was tapped level with markings on the outside wall of the column with a spring-loaded stick terminated by an inverted rubber stopper. This technique insured fairly uniform packing and very little migration of the fine material towards the bottom of each section. For those columns that were to contain stones, each increment was filled with the relative proportions of fine and coarse materials shown in Table 1. This permitted reconstitution of the field soil in a reproducible, although somewhat arbitrarily uniform, manner.

Hydraulic conductivity was measured on these 30 cm length columns under isothermal conditions (23 ± 1 C) by

Table 1. Fractionation analysis of fine and coarse fragments of the soils studied

Soil series	Origin	Texture	Particle size analysis % of <2 mm fraction			Coarse fragments % on total weight basis			
			Clay <2 μ	Silt 2-50 μ	Sand >50 μ	Fines <2 mm	Stones 2-4 mm	Stones 4-9.5 mm	Stones >9.5 mm
"Rock Valley"[1,2]	Mercury, Nevada	Gravelly loamy sand	4.2	18.4	77.4	61.9	4.8	9.8	23.5
Tubac[3]	Silverbell, Arizona	Gravelly sandy loam	13.2	18.3	68.5	73.1	8.9	8.6	9.5
Rillito[3]	Silverbell, Arizona	Gravelly sandy loam	12.3	12.4	75.3	66.2	10.9	9.0	13.9

[1]/not a soil series name

[2]/samples taken from 0-50 cm depth

[3]/samples taken from 0-25 cm depth

Table 2. Some physical properties of the soils studied

Soil series	Bulk densities in $g\ cm^{-3}$ Field ρ_T	Stones ρ_s	Soil ρ_b	EC_e $mmhos\ cm^{-1}$	Saturation % by weight of <2 mm fraction
"Rock Valley"	1.79	2.59	1.50	0.66	28.2
Tubac	1.86	2.60	1.68	0.61	20.5
Rillito	1.88	2.56	1.65	0.34	22.4

the transient outflow method of Weeks and Richards (1967). The equipment and procedures in this study were similar to those described by Weeks and Richards, except for some changes that were necessary to extend the method to approximately -50 bars (the psychrometer range). Details were given in Letey et al. (1973). Soil water diffusivity $D(\Theta)$ was obtained from:

$$D(\Theta) = K(h) \frac{\partial h}{\partial \Theta} \qquad (3)$$

where K (h) is the hydraulic conductivity function of soil water pressure h, and Θ is the volumetric water content (DSCODES A3UST01, 02 and 03).

It was estimated that for these soils the osmotic pressure component, as measured by soil psychrometers, was negligible since the electrical conductivities of the saturated extracts EC_e were low (Table 1). The following example is presented to illustrate this point. For these soils, the volumetric water content is approximately 40 and 6.5% at saturation and at -50 bars soil water pressure, respectively. This represents a six-fold concentration of the soil solution. Assuming all solutes remain in solution -- which is probably overstating our case -- an EC_e of 0.6 $mmhos\ cm^{-1}$ would result in only -1.3 bars osmotic pressure for -50 bars total soil water pressure measured by a soil psychrometer. This value is almost within the precision of soil psychrometers.

The same procedure was followed for columns containing stones as for those that did not. Water contents were calculated on a total volume basis, including stones when present. Since stones hold less water per unit volume than the soil fraction, this procedure resulted in an under-estimation of the actual water content in the soil portion of stony samples. The computer program used to calculate hydraulic conductivity requires that the column cross-sectional area and its length be input values. Volume is obtained by multiplication of these two variables, while volumetric water content is obtained by dividing the total volume of water at each time by this volume. Thus, to correct water contents of non-stony samples for the presence of stones, it was sufficient to rerun the data using a column cross-sectional area increased in proportion to the volume of total soil occupied by stones. The volume ratio of stones to total soil was computed from:

$$\text{volume ratio of stones} = \frac{(1 - W)}{\varrho_s} \varrho_T \qquad (4)$$

where W is the weight ratio of the <2 mm fraction to total soil, ϱ_T and ϱ_s are the measured field bulk density (including stones) and the stone density, respectively. The required increase in column cross-sectional area is achieved simply by dividing the actual cross-sectional area (81.7 cm^2) by 1 - volume ratio of stones. In applying such a correction, the area available for flow is likewise increased. Thus, for an equivalent amount of water passing any plane in a non-stony column the flow velocity is reduced, which in turn results in decreased hydraulic conductivity values. This decrease is included in the calculation of hydraulic conductivity of a stony sample, because the total cross-sectional area of the column is used.

Samples of the "Rock Valley" soil were run in triplicate without stones and in duplicate with stones. No replication was made for the Tubac and Rillito series based on the good agreement observed between replicates of the "Rock Valley" soil. In addition, one column of the latter soil containing only the largest size fraction of stones (>9.51 mm) was also run.

Thermal Moisture Diffusivity under Steady-state Conditions

The steady-state movement of soil moisture, neglecting gravity and the osmotic pressure gradient component of flow, is described by:

$$J_w = -K(h) \frac{\partial h}{\partial x} - L_{wq} \frac{\partial T}{\partial x} \qquad (5)$$

where J_w is the net flux of soil water (liquid and vapor), K(h) is the hydraulic conductivity function of soil water pressure h, L_{wq} is the thermal moisture diffusivity, and T and x are the temperature and horizontal distance respectively.

In a closed system, steady-state conditions are met when the value of the net flux of water J_w is zero. However, this does not mean that no water is flowing. Moisture tends to move as a result of a temperature gradient from warm regions to colder ones usually in the vapor phase. The water content in the colder areas increases by condensation of vapor, thus creating a soil water pressure gradient that will tend to move liquid water in the opposite direction. When these two opposing fluxes are equal, steady-state conditions are met.

Duplicate soil columns, consisting of a piece of acrylic plastic tubing 7.6 cm in diameter and 20 cm in length, were filled with air dry soil. Uniform packing was achieved by filling a 2-cm section of the column at a time and tapping the material even with markings made on the outside wall of the plastic container. Sufficient water was added to each container to bring the <2 mm soil to water contents of 12.5% by volume for "Rock Valley" and 20% for Tubac and Rillito. The soil columns were left to equilibrate for 5 days.

To insure good thermal contact with the soil, bronze plates were fitted to either end of a pair of columns and held in place with silicone seal. At the hot end, a thermoelectric heating pad was inserted between the bronze plates and a piece of masonite, which were bolted to ¾-inch plywood. To control the temperature at the hot end of the column, the heating pad leads were connected to a variable transformer. At the cold end, the bronze plates were attached to a water reservoir connected to the circulating pump of a constant temperature bath. The temperature of the bath was maintained at 21 C throughout the experiment. The paired columns were then placed inside an insulation box and surrounded with coarse vermiculite. Construction details of the equipment are shown in Figure 1.

Copper-constantan thermocouples were inserted at distances of 0, 4, 8, 12, 16, and 20 cm from the cold end of the column through small drill holes. The thermocouple lead wires were then sealed into the holes with non-hardening bathroom seal to facilitate their removal at the end of a run. The thermocouple outputs were recorded on a 12-channel chart recorder (Leeds and Northrup, model Speedomax H) and an ice bath was used as a reference junction.

After another equilibration period of a day or two, the heating pad was activated and temperatures were monitored at periodic intervals. From 5 to 7 days elapsed before the thermocouple outputs at each location along the columns became constant. After several more days, steady-state conditions were assumed to prevail. The apparatus was dismantled and the soil columns were cut into 2-cm sections using a soil core extractor. The sections were placed in covered cans, weighed, oven-dried overnight, and reweighed to determine final water contents and bulk densities. The initial, uniform water content of the column was also calculated by dividing the total loss of water upon drying by the net dry weight of soil. Volumetric water contents were obtained by multiplying the latter values by the bulk density.

The coefficient L_{wq} was calculated for each centimeter along the column from equation (5) in which the net flux J_w is now zero. We obtain:

$$L_{wq} = -K(h)\,\frac{\partial h}{\partial x} \bigg/ \frac{\partial T}{\partial x}\,. \tag{6}$$

The solution to equation (6) presents several problems. The only measured term is the temperature gradient. Since water contents are also measured, the soil water pressure h could be estimated from the soil moisture characteristic curves obtained previously during isothermal experiments. Likewise, hydraulic conductivity K(h) could be taken from the results of the same experiments, providing the values were corrected for temperature. Haridasan and Jensen (1972) have shown that the increase in hydraulic conductivity at a given water content due to a rise in temperature was almost entirely accounted for by the decrease in viscosity of water. The temperature dependence of the Θ(h) function was not known. Although soil water pressure is theoretically directly proportional to the surface tension, its temperature dependence cannot be entirely explained by the temperature dependence of surface

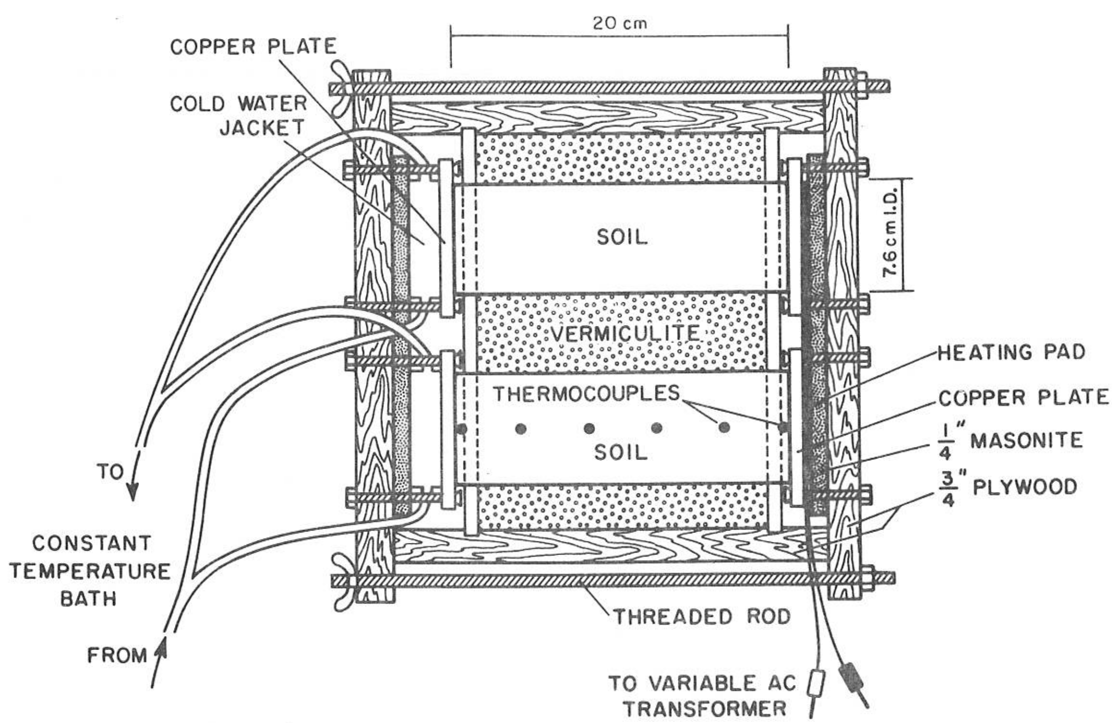

Figure 1. Construction details of equipment used in the steady state study of thermally induced flow.

tension (Haridasan and Jensen, 1972; Jury, 1973).

In order to gain more precision in the calculation of L_{wq}, equation (6) was written in the equivalent form:

$$L_{wq} = -D(\Theta) \quad \frac{\partial \Theta}{\partial x} \bigg/ \frac{\partial T}{\partial x} \qquad (7)$$

where the soil water diffusivity $D(\Theta)$ has replaced $K(h)$ $\partial h/\partial \Theta$. We now have an extra measured term; namely, the volumetric water content gradient. The moisture diffusivity coefficients, as obtained under isothermal conditions (at 23 C) in the previous experiments, but corrected for temperature, were used for the $D(\Theta)$ values. The variation of this coefficient with the temperature was satisfactorily described by the temperature dependence of the ratio of surface tension to viscosity of water (Jackson, 1963). Soil water diffusivities were thus corrected using:

$$D(\Theta)_T = D(\Theta)_o \frac{\sigma_T \, \eta_o}{\sigma_o \, \eta_T} \qquad (8)$$

where σ is the surface tension of water in contact with air (dynes cm^{-2}), η is the viscosity of water (centipoise), and the subscripts o and T refer to the reference and measured temperatures, respectively. Treatment of the data according to equations (7) and (8) still assumes proportionality between soil water pressure and surface tension.

The volumetric water content and temperature gradients were calculated with the IBM 360 Data Processing System using a numerical differention subroutine written in Fortran IV. The subroutine included smoothing and wild point replacement procedures. Comparison between raw data and smooth data, after wildpointing, revealed only very small differences and the method was found to be much more precise than either graphical or simple numerical methods that could be carried out on a small desk calculator.

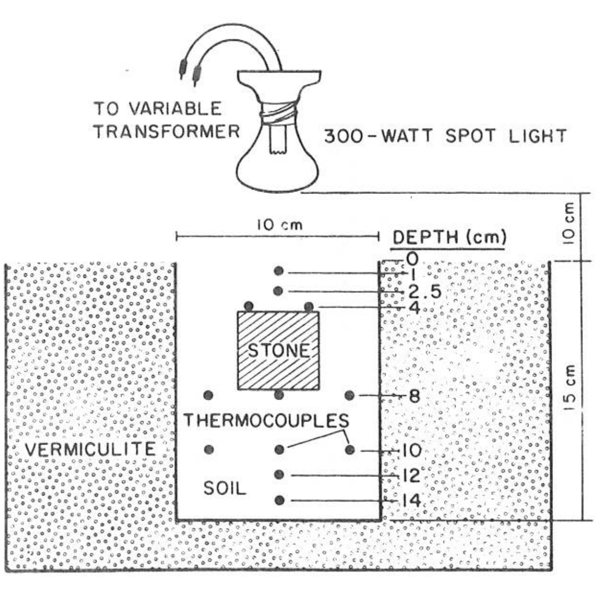

Figure 2. Diagram showing locations of copper-constantan thermocouples in buried stone experiment.

The same experimental procedure was followed for columns both with and without stones. Water contents were based on the total volume of the column (A3UST04).

Diurnal Temperature Distributions Under Stones

In an attempt to find out whether there existed a large enough temperature difference for condensation to occur on the undersides of stones, experiments were set up to measure the temperature profiles underneath and in the vicinity of stones.

A plastic container 10 cm in diameter by 15 cm in length was filled with air dry "Rock Valley" soil. A stone, 5 x 4 x 4 cm (approximately) was buried 4 cm from the surface during the packing process. Copper-constantan thermocouples were inserted at several locations as shown in Figure 2. The soil container was placed inside a box filled with vermiculite to insure lateral thermal insulation. Heat was supplied by a 300-watt spotlight held at 10 cm from the soil surface. The lamp was connected to a variable transformer so that the soil surface temperature could be controlled. The thermocouple outputs were recorded on a 12-channel chart recorder (Leeds and Northrup, model Speedomax H) and an ice bath served as a reference junction.

Heat was supplied for 9.5 hr, then the lamp was turned off and the soil was allowed to cool overnight. A second run was made after addition of sufficient water to bring the soil to a water content of 10% by volume.

In the second experiment, a plastic tub, 56 cm in diameter and 15 cm high, was filled with air dry "Rock Valley" soil. A large granitic stone 12 cm high, 30 cm long and 19 cm wide was placed on its surface. A diagram (Fig. 3) shows the thermocouple locations. Four 300-watt lamps provided heat. Because of the large dimensions of the tub, lateral thermal insulation was estimated not to be necessary. Two runs were made similar to the first experiment and temperatures were monitored on a continuous basis (A3UST05).

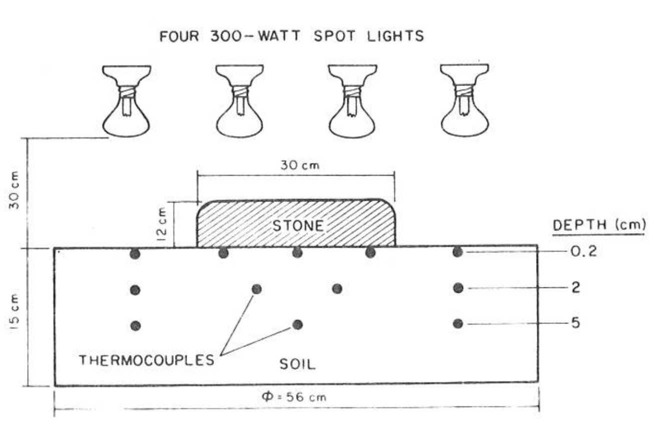

Figure 3. Diagram showing locations of copper-constantan thermocouples in surface stone experiment.

RESULTS AND DISCUSSION

The following data have been or will soon be submitted in machine-readable form to the Desert Biome central data bank:

1. Soil water pressure, volumetric water content, hydraulic conductivity and soil water diffusivity at three distances along the soil columns for "Rock Valley" soil (A3UST01, 02).
2. Same as above for Tubac and Rillito soils (A3UST03 to be submitted).
3. Temperatures, water contents, and thermal moisture diffusivities for "Rock Valley" soil (A3UT04 to be submitted).
4. Temperatures at selected depths under and around stones submitted to a diurnal heat wave (A3UST05 to be submitted).

An interpretive summary of the results of research conducted in 1973 follows.

Hydraulic Conductivity

The volumetric water content-soil water pressure relationship for each soil as calculated from the transient outflow method during desorption are presented as solid and dashed lines in Figure 4. Data points correspond to water contents determined gravimetrically at the end of a run for which the soil water pressure was known. The generally good agreement between experimental points and calculated curves leads to the conclusion that the transient outflow method is appropriate to relatively dry soil systems.

Unsaturated conductivity values calculated for stony samples of "Rock Valley" and Tubac soils over the soil water pressure range -0.05 to -50 bars are shown in Figure 5. These are typical examples of the results obtained in these

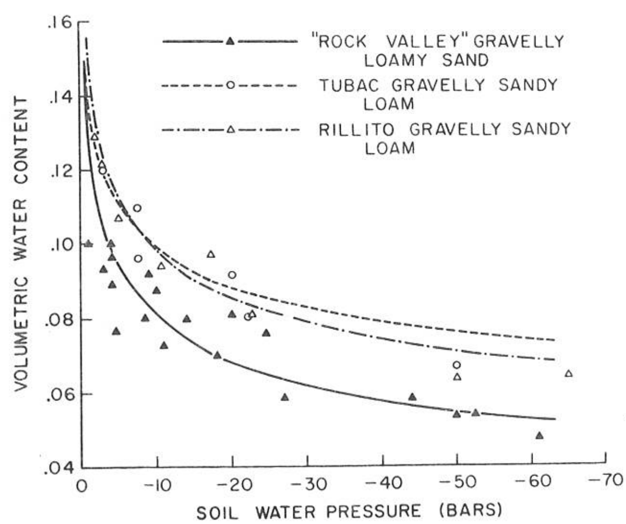

Figure 4. Moisture characteristic curves for "Rock Valley" gravelly loamy sand, Tubac gravelly sandy loam, and Rillito gravelly sandy loam during desorption (A3UST02 and ST03).

soils. Results for Rillito were very similar to those for Tubac. The experimental points for K as a function of h are given for three positions along the column corresponding to distances of 2.5, 5 and 10 cm from the drying end. Some replication is obtained with the transient outflow method because similar values of h occur at different positions along the column at different times. Since the experimental points for different positions but similar h values are within experimental error of each other, the extension of the transient outflow method to soil water pressures beyond the tensiometer range is further justified.

Plots similar to those shown in Figure 5 were obtained for each soil column Distinguishing between non-stony and stony cases, best by-eye fit curves were drawn through all data points and replicates were averaged in the case of "Rock Valley" soil (Figs. 6, 7 and 8). It is immediately apparent that there is little difference between stony and non-stony samples when hydraulic conductivity is expressed as a function of soil water pressure. As shown in Figure 3, even varying the weight ratio of stones to total soil from as much as 38.1 to 23.5% had no effect on the relationship between K and h.

In the case of Tubac soil (Fig. 7), agreement between stony and non-stony samples is not so good, especially at the higher soil water pressures (greater than -1 bar). In the dryer range, however, the curves tend to be quite similar. Tensiometers or thermocouple psychrometers measure the soil water pressure of the soil fraction only, since they do not come in contact with stones. The soil water pressure values obtained should therefore not be affected by the presence of stones, as was indeed found with "Rock Valley" and Rillito (Figs. 6 and 8). The poor agreement observed with Tubac could possibly be the result of experimental error. Further replication is warranted.

When unsaturated hydraulic conductivity is expressed as a function of volumetric water content, distinct curves are obtained for stony and non-stony samples (Figs. 9, 10 and 11). For the "Rock Valley" soil in Figure 9, each line represents the average of replicates. For similar K values, water contents are always higher when stones are excluded because, in the transient outflow method, volumetric water contents are calculated on the basis of the total volume of the column. In other words, stones are considered to hold as much water as the soil portion does. But stones hold much less water than soil at comparable soil water pressures, so that the water contents of the soil fraction computed for soil samples containing stones are underestimated. With increasing water content, the underestimation also increases.

It is not implied that water contents for stony soil based on total volume are erroneous. These are the values that a field worker would obtain by direct gravimetric sampling or from neutron probe readings. Therefore, on the assumption that the stony fraction did not absorb water, hydraulic conductivity curves for samples without stones were redrawn by correcting water contents as though stones were present. For "Rock Valley", and especially Rillito (Figs. 9 and 11), the correction appears to be appropriate. For

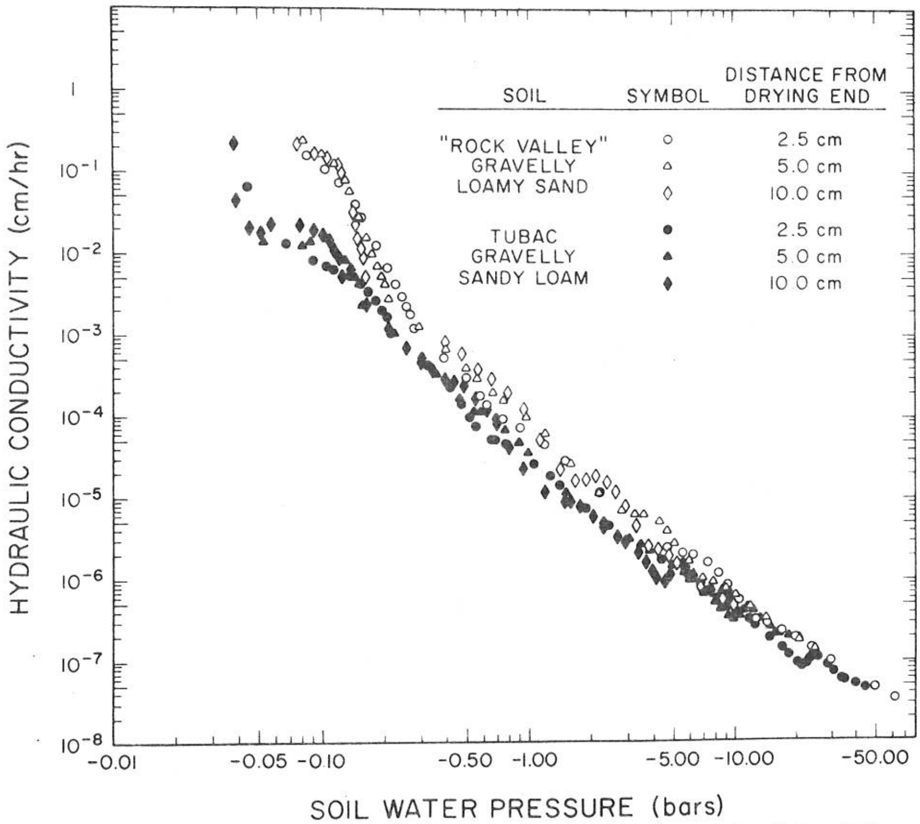

Figure 5. Experimental points given for unsaturated hydraulic conductivity at three locations and plotted as a function of soil water pressure for stony samples of "Rock Valley" and Tubac soils (A3UST02 and ST03).

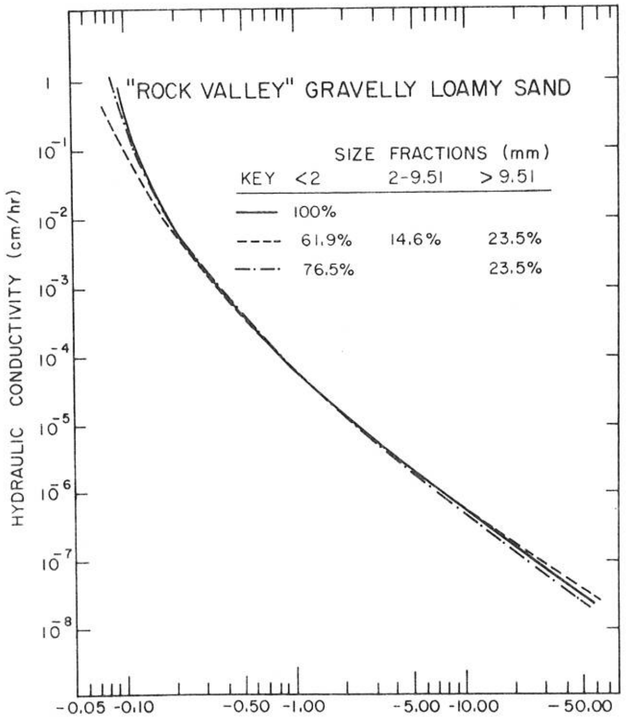

Figure 6. Hydraulic conductivity as a function of soil water pressure for stony and non-stony samples of "Rock Valley" gravelly loamy sand. Curves represent the average of replicates (A3UST02).

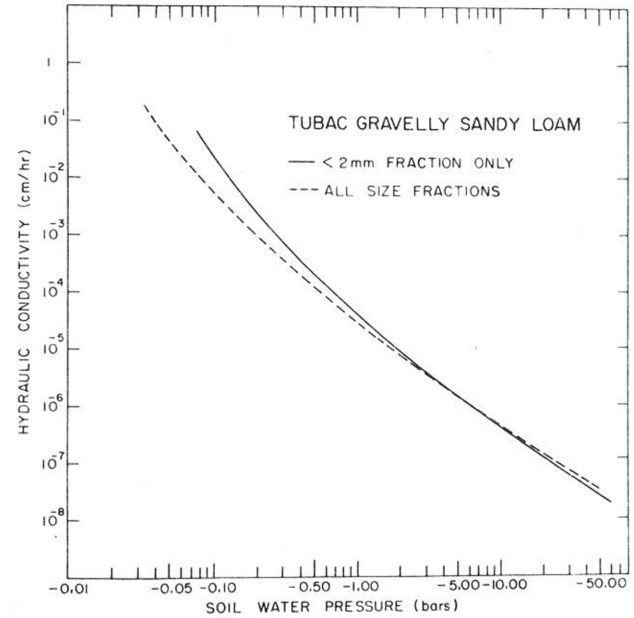

Figure 7. Hydraulic conductivity as a function of soil water pressure for stony and non-stony samples of Tubac gravelly sandy loam (A3UST03).

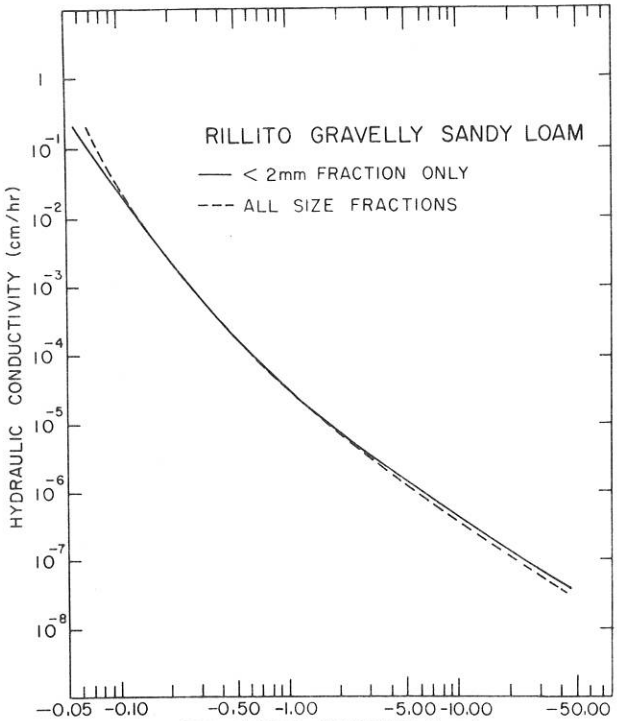

Figure 8. Hydraulic conductivity as a function of soil water pressure for stony and non-stony samples of Rillito gravelly sandy loam (A3UST03).

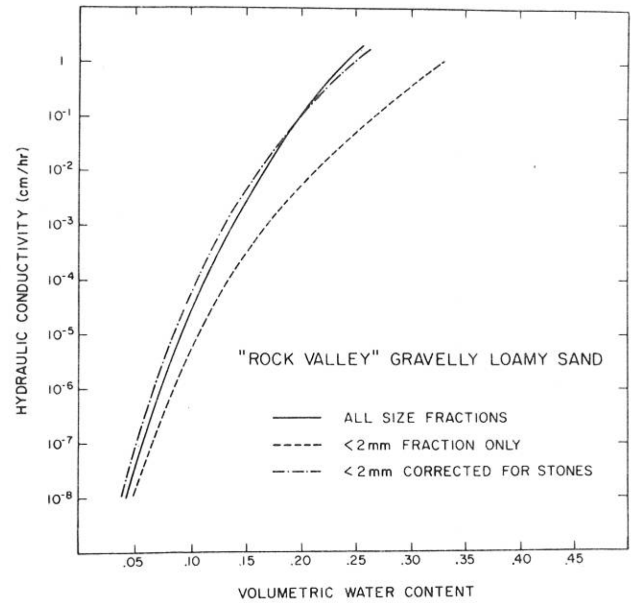

Figure 9. Relation between hydraulic conductivity and volumetric water content for "Rock Valley" gravelly loamy sand with and without stones. Curves represent the average of replicates (A3UST02).

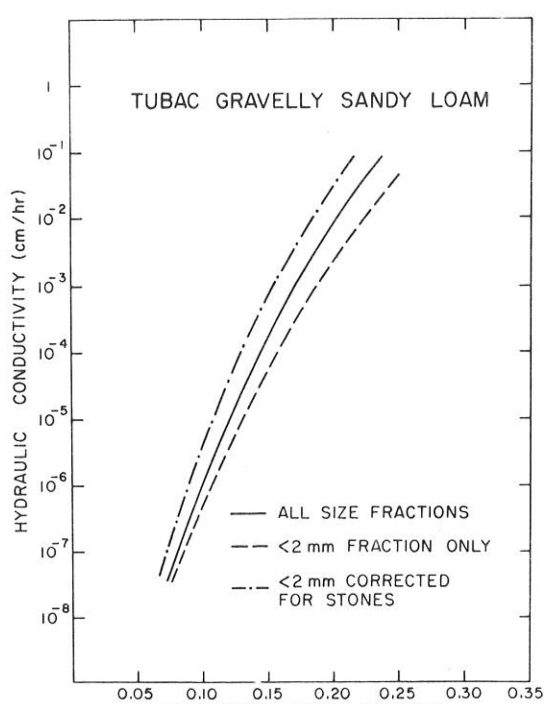

Figure 10. Relation between hydraulic conductivity and volumetric water content for Tubac gravelly sandy loam with and without stones (A3UST03).

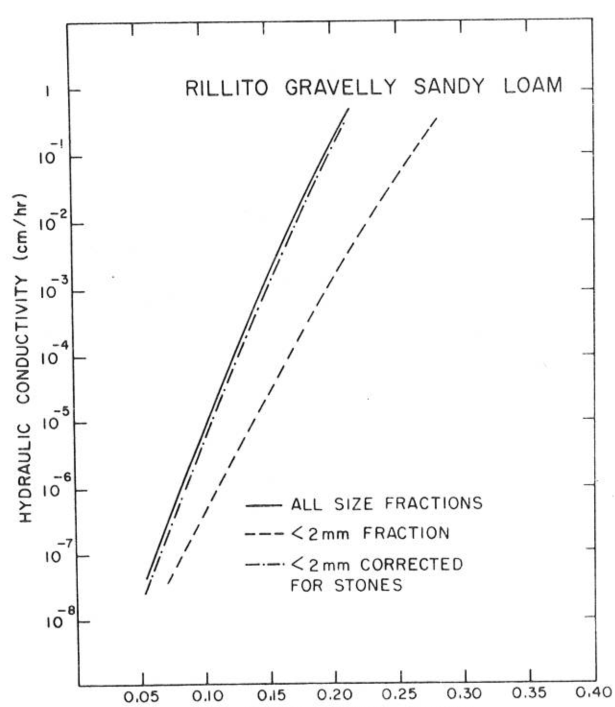

Figure 11. Relation between hydraulic conductivity and volumetric water content for Rillito gravelly sandy loam with and without stones (A3UST03).

Tubac (Fig. 10) the $K(\theta)$ curve was overcorrected.

It is difficult to explain the behavior of this latter soil. More replication might have provided better agreement. Recall that the fit between stone-free and stony samples of this soil was not too good when K was expressed as a function of h. While there is good reason to believe it should have been better, and therefore also K versus θ, the validity of the assumption that stones do not absorb water at all may be questionable. For example, Reinhart (1961) estimated that stones in the Fernow Forest soils of West Virginia, which are derived from sandstone and shale, usually had a gravimetric water content of 6.5%. Furthermore, moisture content of stones may vary appreciably with soil water content and thus the correction used would be different at different moisture levels. Coile (1953) prepared a simple regression graph relating moisture content of stone to uncorrected soil water content for soils derived from various and differing parent materials. The moisture content of the stones was significantly less than that of the finer material. Unfortunately, his data are restricted to relatively wet moisture levels. The regression line he drew through the data points indicates that at zero moisture content of total soil (stones included), stones would have a moisture content close to 4% by weight. Hence, Coile's correction is useless at the water contents encountered in this study.

Examination of Figure 10 reveals a greater overcorrection at the higher water contents than at low water contents. It appears that the stone portion of the soil did in fact absorb increasing amounts of water with increasing soil moisture contents. This soil is derived from andesite, basalt, granite, and quartzite parent materials. The assumption that stones do not hold water is probably only correct for quartz gravel but is not necessarily true for rock composed of other minerals in various stages of weathering.

At saturation, separate stone samples were found to hold 4.1, 7.6 and 9.0% of water by volume for "Rock Valley", Tubac and Rillito, respectively. Rillito gave the best corrected results, but also had the highest stone water content at saturation, which would appear to be contradictory. So until measurements are made of the moisture content of stones with varying soil moisture content, corrections must be made on the basis that small stone does not contain appreciable amounts of water. It is realized that this approach may not always be fully justified. In practice, a simple correction of water contents and of the area available for flow would be sufficient in most cases to account for the presence of stones. If the amount of water held by stones were known as a function of soil water content, then a progressive correction could be applied to values obtained on columns without stones.

On the other hand, if K is expressed as a function of h, as in Figures 6, 7 and 8, hydraulic conductivity values obtained by the transient outflow method on columns of soil with stones excluded provide a good estimate of the values that would be obtained if stones were present, without any special treatment of the data.

Thermal Moisture Diffusivity

The initial water contents of Tubac and Rillito soil columns were nearly 20% by volume in the <2 mm fraction. The water content distributions at the end of the experiments had not changed significantly from the initial values. At these high moisture levels, hydraulic conductivity is also large. As a consequence, only a very slight pressure head gradient will be required to move water. As soon as a small amount of water moves from the hot to the cold end of a column as a result of the imposed temperature gradient, it will move right back under the opposing soil water pressure gradient. Because water cannot escape a sealed column, the end result is a steady-state water content distribution equal to the initial uniform water content. For these reasons, it was impossible to evaluate the coefficient L_{wq} for these soils as a function of water content. But it was estimated that, for $\theta = 0.20$, L_{wq} would have to be in the range of 0.007 to 0.010 $cm^2\ hr^{-1}\ deg\ C^{-1}$.

"Rock Valley" soil samples were run at the considerably lower initial water content of 12.5% in the <2 mm fraction. The steady-state temperature and moisture content distributions as a function of distance from the cold end are shown for individual columns in Figure 12. It is immediately apparent that the temperature distributions in paired, duplicate columns were not identical. The bronze plate (see Fig. 1) farthest away from the heating pad lead wires (dashed lines in Fig. 12) received less heat. The temperature reached at the hot end of these columns was 5 C lower. As a result, the warmer columns had a lower water content at the warm end and a higher water content at the cool end, as would be expected. Because the moisture distributions in duplicate columns were different, it was not feasible to average them. The calculations required to obtain L_{wq} were thus performed for individual columns.

The data points in Figure 12 correspond to the measured water contents of each 2-cm section, while the lines represent the smoothed values obtained from the numerical differentiation subroutine. In the absence of stones, smoothed data agree closely with raw data. When stones were present (bottom of Fig. 12), raw data do not fall on the smooth curve.

The scatter of measured values points out the difficulties of sampling stony soils. When the columns were sectioned at 2-cm intervals, sometimes a stone would come into contact with the cutting implement. There was no alternative but to include the stone with the subsample being sectioned, although part of it belonged to the next section. Thus, the volume and air dry weight of subsamples varied, resulting in only approximate estimates of the water contents in each section. Non-destructive determination of water contents by gamma-ray attenuation would most likely impose similar problems due to the very narrow bands that are scanned by such instruments.

The coefficient L_{wq}, which is termed thermal moisture diffusivity by some workers, was calculated from the curves of Figure 10 except where the water content gradient was positive as in the center of the stony columns. Results for duplicate columns were plotted together as a function of volumetric water content (Fig. 13). Water content values were based on total volume; they were not corrected for the presence of stones.

Because L_{wq} values are widely scattered, a curve was not drawn through the data points. The results do suggest, nonetheless, a decreasing relationship with decreasing water content. This is in contrast to the findings of Letey (1968) who analyzed the work reported in several published articles. On the basis of available data, he found that L_{wq} remained fairly constant over a wide range of water contents when corrected to a common temperature of 22 C. The correction was applied assuming that all thermally driven moisture movement occurred in the vapor phase according to principles set forth by Cary (1965). The theory of Philip and de Vries (1957) predicts thermal vapor diffusivities to be fairly constant as water content changes, whereas thermal liquid diffusivities would decrease with water content. Examination of equation (6) reveals that as the water content decreases, hydraulic conductivity decreases more rapidly than the absolute value of the soil water pressure gradient increases. The temperature gradient varies little in most soil systems. The net result would cause L_{wq} to decrease also and at a faster rate at high than at low water contents.

Indeed, the coefficient L_{wq} in Figure 13 appears to start leveling off at about $\theta = 0.10$. It may be surmised that for this soil, at water contents above about 0.10, thermally induced moisture flow probably takes place in the liquid phase. This is the region where L_{wq} decreases most rapidly. Jury (1973) reported L_{wq} values for Plainfield sand at high water contents where flow was most certainly in the liquid phase. L_{wq} decreased very rapidly, almost logarithmically, with water content.

At water contents below about 0.10, vapor movement becomes increasingly important. For the purpose of comparison, theoretical thermal vapor diffusivities, calculated according to Philip and de Vries (1957), are shown in Figure 13. These values increase slowly with decreasing water content reaching a plateau of about 0.001 $cm^2\ hr^{-1}\ deg^{-1}$ at $\theta = 0.05$. At moisture levels in the vicinity of 0.07, thermally driven liquid flow has probably ceased and all the thermally induced moisture movement is in the vapor phase. Visual extrapolation of the data points in Figure 13 tends to confirm this hypothesis.

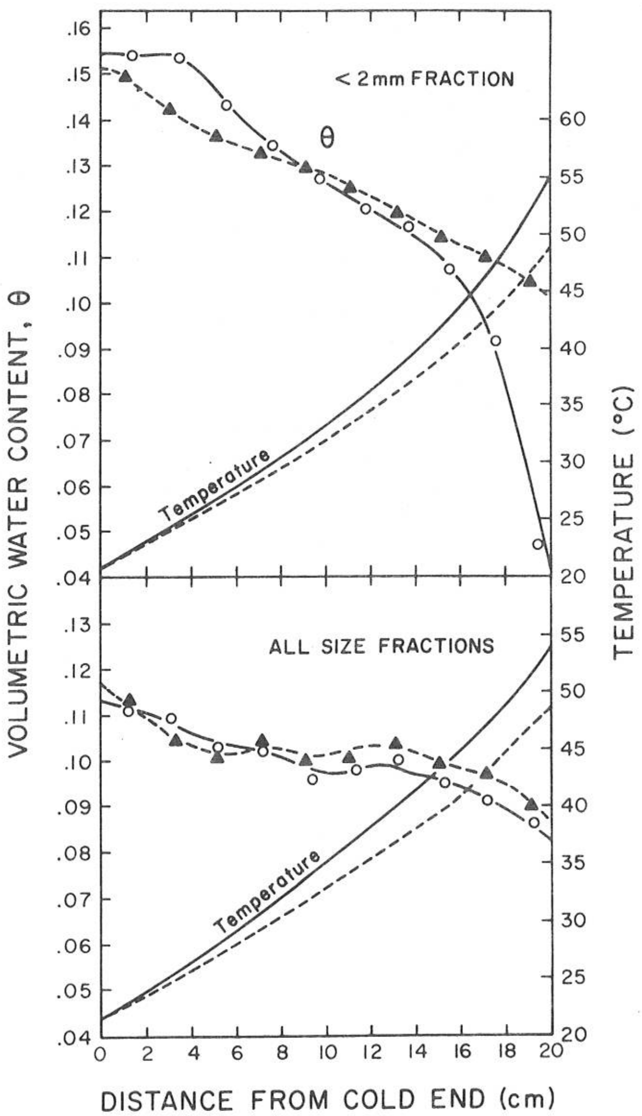

Figure 12. Volumetric water content and temperature distributions in duplicate columns of "Rock Valley" gravelly loamy sand under steady state conditions (A3UST04).

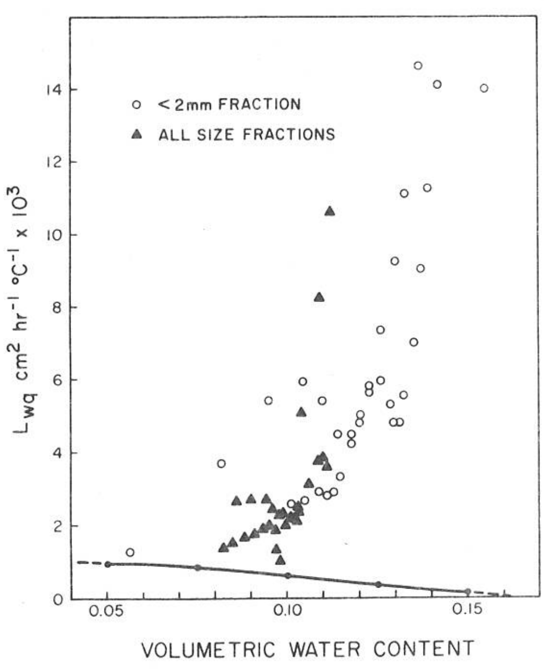

Figure 13. Values of the coefficient L_{wq} associated with moisture movement under temperature gradients for "Rock Valley" gravelly loamy sand (A3UST04).

Although the coefficient L_{wq} is temperature dependent, values of L_{wq} were not corrected to an average common temperature as Letey (1968) suggested because use of his correction procedure assumes solely a vapor transfer process. Data of Cary and Taylor (1962a, b), as analyzed by Letey, indicate that the coefficient affecting water flow due to temperature gradients in the liquid phase would be more temperature dependent than in the vapor phase. The magnitude of the correction to be applied is unknown. However, the decrease of L_{wq} with θ might be less sharp were L_{wq} corrected for temperature differences.

No attempt was made to correct the results obtained on stone-free samples to those of stony samples. The relations between L_{wq} and θ pertaining to each type of sample were not deemed sufficiently defined to permit a generalized statement. Let it be noted, though, that contrary to hydraulic conductivity the necessary correction would not be as simple. Water contents would have to first be corrected as was done in the preceding section. To the new set of water content values would correspond new moisture diffusivities which would then have to be adjusted for temperature according to equation (8). Correcting the water contents of non-stony samples as though stones were present would lower their value. Therefore, $D(\theta)$ would also be lowered. If both variables decreased in such a fashion as to cause L_{wq} to lie further down but on the same curve, no correction would actually be necessary. Additional and more precise data are required to verify this possibility.

Diurnal Temperature Distributions Under Stones

The temperature distributions in the case of a buried stone are shown in Figures 14 and 15 for air dry soil and soil brought to a moisture content of 10% by volume, respectively. Temperatures during the cooling period were not monitored with the dry soil. After 9.5 hr of heating, the maximum range in temperatures achieved at the 1-cm depth with air dry soil was 35 C, while at 14 cm below the soil surface, the difference in temperature was only 8 C (Fig. 14). With moist soil, the maximum fluctuation at 1 cm was reduced to 25 C, while at the lower depth is was increased to 10 C for a comparable heating period of 9.5 hr (Fig. 15). The decrease in maximum amplitude for moist soil at or near the soil surface is attributed to the cooling effect of water evaporating into the atmosphere. The increase in maximum amplitude at the lowest depth is due to the higher thermal conductivity of a moist soil over a dry one and to the transport of sensible and latent heat as the soil water is redistributed within the profile. In the moist soil, cooling was very rapid at the surface and the temperature gradient reversed directions after only 1.5 hr, resulting in lower temperatures at the surface than at depth, as was expected.

In neither case -- air dry or moist soil -- was it possible to distinguish a temperature difference at the 8 cm depth between the lower surface of the buried stone and the adjacent soil. Presumably, heat is transferred through the soil around the stone, warming or cooling the under surface of the stone at the same rate as the soil mass itself. It is therefore unlikely that water vapor could condense beneath buried stones as the sole result of differing thermal properties between stones and soil. Subterranean dew might be formed, however, in the first few centimeters of the profile simply because of the upward water vapor movement during cloudless nights. This phenomenon would not be restricted to the undersurfaces of stones but would take place in the adjacent soil as well.

When a large stone was placed on the surface of air dry soil (Fig. 16), the daytime temperatures directly beneath the stone were lower than at the soil surface. This lag in temperature rise was, however, overcome at the end of the heating period. The maximum difference recorded was 7 C towards the middle of the heating period. The same was true at a depth of 2 cm but to a lesser degree. The largest

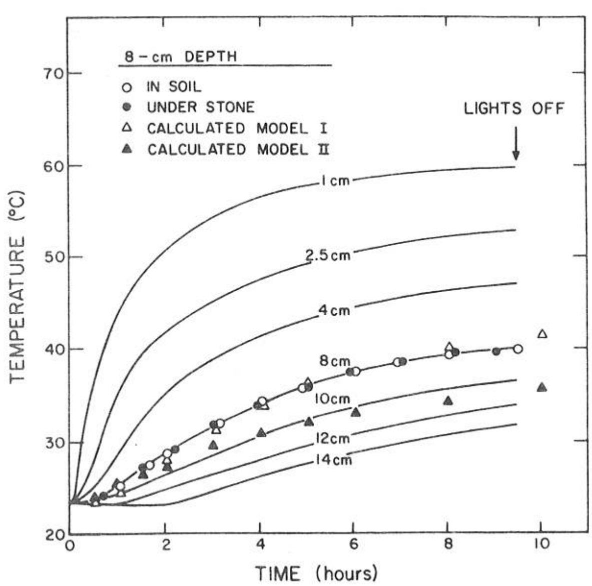

Figure 14. Variation of temperature with time at various depths for a buried stone in air dry soil (heating only); A3UST05.

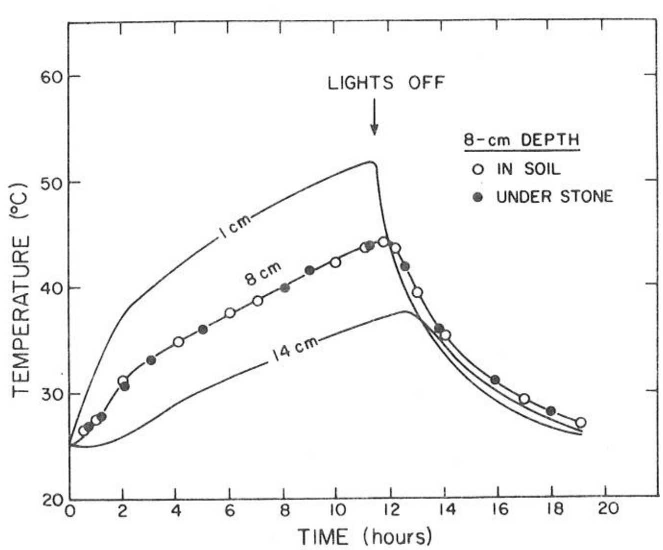

Figure 15. Diurnal variation of temperature at selected depths for a buried stone in moist soil (A3UST05).

temperature difference here was 5 C. At 5 cm below the soil surface, no temperature difference was recorded between thermocouples placed under the stones and those inserted in the soil away from it. At night, during the cooling period, the trend was reversed. Temperatures under the stones were higher than those at similar depths in the adjacent soil. At the end of the cooling period, temperatures were similar throughout the profile.

During the early part of the day, the lower temperatures below surface stones will result in a vapor density gradient causing water vapor to flow from the surrounding area toward the undersurface of the stone. The difference in temperature (7 C) might be large enough to cause vapor to condense if conditions of temperature and relative humidity in the soil are right. In any event, the presence of large surface stones would tend to reduce evaporation by obstructing diffusion of the vapor into the atmosphere.

Stark and Love (1969) reported occurrences of condensation on the undersides of "...rocks near the surface. These rocks appear to heat severely during the day but radiate freely at night and thus become cooler than the surrounding soil mass...Rocks which were dry in the p.m. would actually be moist to wet on the undersides in the early a.m." These workers did not explicitly measure temperatures under stones, however, nor do they mention at what time their observations were made. It is quite possible that "early a.m." could be several hours after sunrise, especially on a summer day. The condensation they noticed could well have accumulated between dawn and the time of their observations.

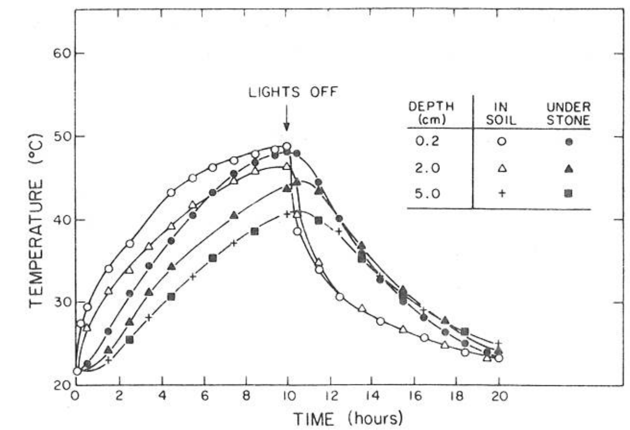

Figure 16. Diurnal variation of temperature at various depths when a large stone is placed on the surface of air dry soil (A3UST05).

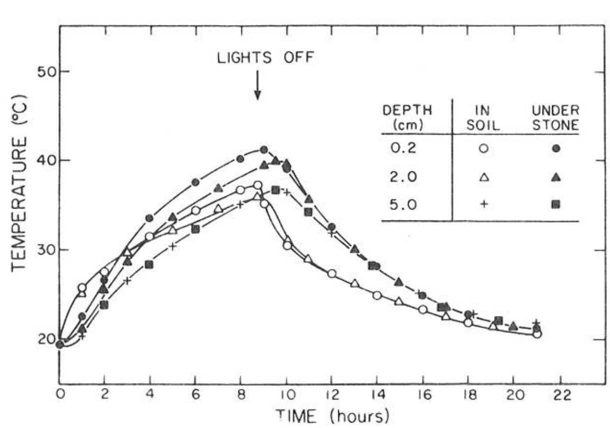

Figure 17. Diurnal variation of temperature at various depths when a large stone is placed on the surface of moist soil (A3UST05).

At night, the gradient of temperature is reversed causing water to move away from under the stones. But during this time, evaporation into the atmosphere is practically reduced to zero because the air temperature is lowered, while the air relative humidity rises concomitantly. Therefore, water that moves away from the stones at night would not be lost in any appreciable amount to the atmosphere. Furthermore, the water content of the upper layers of soil may be increased at night by the ascending soil moisture and also by condensation of atmospheric dew due to nightly radiation.

The combination of the processes during the day and night could result in a net accumulation of water under surface stones in an arid environment. The results of these simple experiments do not imply that this does happen, but merely that it could. The question of whether the amount of water that accumulates in this fashion is physiologically significant to small animals and plants remains to be answered.

Under moist conditions, temperatures under a surface stone (Fig. 17) were higher than in the surrounding soil at all times of the day. The effect was still felt at the 2 cm depth, but not at 5 cm where temperatures measured under the stone and away from it were identical. The results are in agreement with those of Lamb and Chapman (1943) and Saini and MacLean (1967) who found higher summer temperatures under stones in moist soils. Temperatures at the soil surface were lower than directly beneath the stone and also lower than in dry soil because of loss of latent heat to the atmosphere. The difference in temperatures between the surface and 5 cm below the surface was lower than in air dry soil (Fig. 16) because of the increase of thermal conductivity in moist soil and the transport of sensible heat with the redistribution of soil water.

EXPECTATIONS

The research conducted in 1973 yielded data on the water transfer properties of Rock Valley and Silverbell soils. Similar data will be collected for the Jornada soils.

Studies with Silverbell soils will be continued, particularly the effect of temperature gradients on soil water flux. The effect of stones on soil water transfer functions will be further investigated.

Diurnal variations of temperature under large surface stones will be monitored *in situ* and the extent to which water may accumulate below them will be determined.

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